Characterizing a Ground Water Basin in a New England Mountain and Valley Terrain

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Abstract

A ground water basin is defined as the volume of subsurface through which ground water flows from the water table to a specified discharge location. Delineating the topographically defined surface water basin and extending it vertically downward does not always define the ground water basin. Instead, a ground water basin is more appropriately delineated by tracking ground water flowpaths with a calibrated, three-dimensional ground water flow model. To determine hydrologic and chemical budgets of the basin, it is also necessary to quantify flow through each hydrogeologic unit in the basin. In particular, partitioning ground water flow through unconsolidated deposits versus bedrock is of significant interest to hillslope hydrologic studies. To address these issues, a model is developed and calibrated to simulate ground water flow through glacial deposits and fractured crystalline bedrock in the vicinity of Mirror Lake, New Hampshire. Tracking of ground water flowpaths suggests that Mirror Lake and its inlet streams drain a ground water recharge area that is about 1.5 times the area of the surface water basin. Calculation of the ground water budget suggests that, of the recharge that enters the Mirror Lake ground water basin, about 40% travels through the basin along flowpaths that stay exclusively in the glacial deposits, and about 60% travels along flowpaths that involve movement in bedrock.

Introduction

In hydrologic and ecosystem investigations of mountainous or hilly terrain, the study unit is typically a surface water drainage basin, or catchment, outlined by the surrounding topographic divide. In many catchment studies, the flow of ground water is assumed to occur in the shallow, unconsolidated material of the basin (e.g., Pearce et al. 1986; McDonnell et al. 1991; Turton et al. 1992; Peters et al. 1995). The assumption of shallow ground water flow beneath the surface water basin implies that ground water does not flow across topographic divides, and flowpaths do not enter the underlying bedrock.

Several recent catchment studies, however, have found evidence of ground water flow across topographic divides and through bedrock. In a study of a forested catchment in Oak Ridge, Tennessee, Genereux et al. (1993) and Mulholland (1993) found that ground water recharge from outside the catchment is a significant contributor to stream flow in the catchment. In a study of a surface water basin in Alberta, Canada, Crowe and Schwartz (1985) found that most ground water discharge to a lake occurs through bedrock, although they pointed out that the overall contribution of ground water inflow to the lake was small in comparison to inflow from surface water and direct precipitation. In the ground water literature, the concept of flow across topographic divides is well recognized in studies of regional-scale flow (e.g., Toth 1962; Freeze and Witherspoon 1967), lake–ground water interactions (Winter 1978), and flow in hummocky terrain of the midwest United States (Swanson et al. 1988). These studies suggest that the primary basis for delineating a ground water basin should be the ground water flowpaths rather than the surface topography, although topography can exert a strong influence on the flowpaths.

In this paper, we define a ground water basin as the volume of subsurface through which ground water flows from the water table to a specified discharge location. We delineate the ground water basin by tracking flowpaths from discharge points back to the water table. To illustrate this approach, we present a study of a ground water basin in a New England mountain and valley terrain consisting of glacial deposits overlying bedrock. In particular, we investigate the three-dimensional extent of the ground water basin, the possibility that ground water might flow across topographic divides, and the partitioning of ground water flow through glacial deposits and bedrock.

The Study Area

The study area, known as the Mirror Lake area, is about 15 km north of Plymouth, New Hampshire (Figure 1). It occupies about 10 km^2, including Mirror Lake, and lies partially within the Hubbard Brook Experimental Forest, a center operated by the U.S. Department of Agriculture Forest Service (1991) for long-term watershed research. The land surface is characterized by steep hillsides with high-gradient streams, and relatively flat river valleys. Land-surface altitude ranges from 180 m above sea level (masl) at the Pemigewasset River to 700 masl at the ridge top near the northwestern corner of the study area. Average precipitation from 1978

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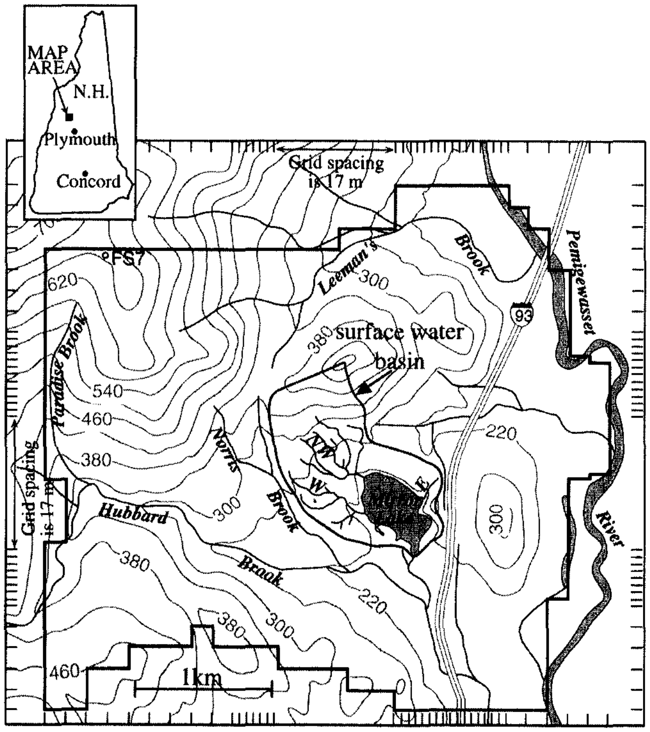

Figure 1. Location map of the Mirror Lake area. Gray contours are land surface elevation, in meters above sea level (masl); thick solid line is lateral boundary of the flow model domain; tick marks on outer box indicate the finite-difference grid discretization.

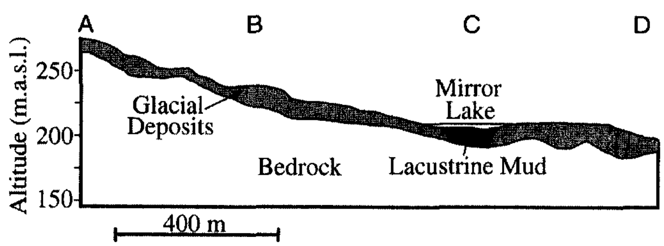

Figure 2. Hydrogeologic section A-B-C-D. (See Figure 3 for location of section line.)

to 1986 was 124 cm year^{-1} (Federer et al. 1990). Three streams, designated as W, NW, and E, flow into Mirror Lake. Together, the lake and the three inlet streams drain a surface water basin that occupies 0.85 km^2 (Figure 1).

Glacial deposits cover much of the land surface in the Mirror Lake area (Figure 2). The glacial deposits consist primarily of till, with local deposits of sand and gravel, and are 0 to 50 m thick. Based on results of slug tests, Winter and Rosenberry (1995) estimated a hydraulic conductivity range of 2.6×10^{-7} to 1.3×10^{-6} m s^{-1} for till, and 1.8×10^{-6} to 2.9×10^{-4} m s^{-1} for sand and gravel. The glacial deposits overlie bedrock that largely consists of a high-grade metamorphic schist with extensive granitic intrusions. Both the schist and granite are intruded by pegmatite dikes, and all three rocks are cut by lesser amounts of lamprophyre, a fine-grained volcanic dike rock. The spatial distribution of rock types is highly complex. Exposed rocks on roadcuts show that the bedrock composition is highly variable over distances of tens of meters. Drill cuttings and video image logs show that there is generally poor correlation of rock types between wells drilled tens of meters apart.

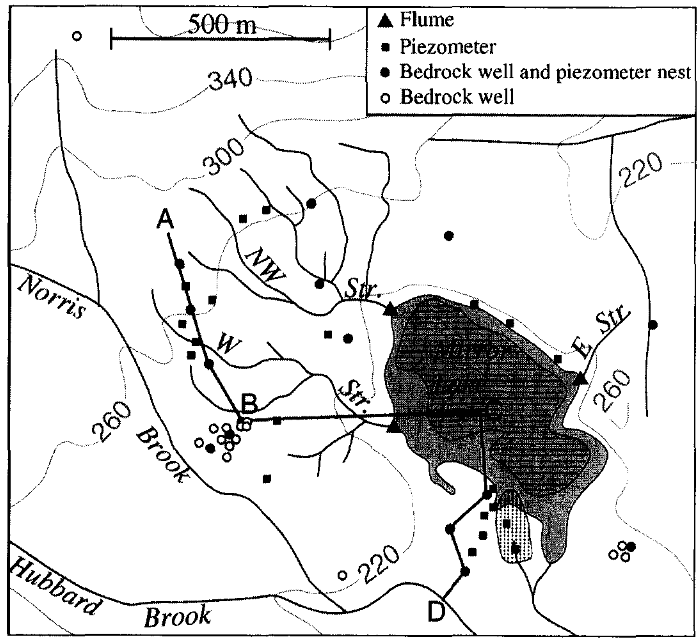

Figure 3. Locations of flumes, piezometers, and bedrock wells except FS7. Striped area within Mirror Lake represents the lacustrine mud, stippled area on south shore of Mirror Lake represents the sand and gravel deposit.

Fractures are the major conduits through which ground water flows through the bedrock. Results of 208 packer tests using 4 to 5 m test intervals in bedrock wells show that bedrock hydraulic conductivity varies over at least six orders of magnitude, from below 1×10^{-10} m s^{-1} (lower limit of measurement) to 5×10^{-5} m s^{-1}. Multiple-well hydraulic tests indicate that the bedrock contains highly conductive fracture clusters, each cluster occupying a near-horizontal volume approximately 1.5 m thick and 20 to 50 m in horizontal extent, and having a hydraulic conductivity of 6×10^{-5} m s^{-1} (Hsieh and Shapiro 1994, 1996). The highly conductive fracture clusters are hydraulically connected to each other through a less conductive fracture network. The multiple-well test results suggest that, over distances of tens of meters, ground water could travel through the highly conductive fracture clusters, but over distances of 100 m or more, ground water must flow through the less conductive fracture network. Based on results of the multiple-well tests, Hsieh and Shapiro (1994) estimated an effective hydraulic conductivity of 2×10^{-7} m s^{-1} for a 100 m size block of bedrock.

Mirror Lake is a kettle lake with a maximum water depth of 11 m and an average water depth of about 6 m (Winter 1984). A layer of lacustrine mud occurs over much of the lake bottom (Figure 3). Hydraulic conductivity of the lacustrine mud is estimated to range from 10^{-10} to 10^{-9} m s^{-1} (Rosenberry and Winter 1993). In the central part of the lake bottom, the lacustrine mud is believed to lie directly over bedrock, while away from the central region, the mud overlies glacial deposits. Along the near-shore lake bottom (littoral zone), the lacustrine mud is very thin to nonexistent. Bedrock crops out in a small area on the east side of the lake.

The ground water system in the Mirror Lake area is recharged by: (1) infiltration from precipitation to the water table; (2) subsurface outseepage of lake water from Mirror Lake; and (3) infiltration of stream water along losing reaches of streams. Infiltration from precipitation is the largest source of recharge. Based on streamflow recession analysis, Mau and Winter (1997) estimated that recharge from precipitation varies from about 19 to 53 cm year^{-1}, or about 15 to 43% of precipitation. Subsurface outseepage from Mirror Lake occurs along a part of the near-shore lake bottom at the south side of the lake, where a deposit of highly permeable sand and gravel allows rapid infiltration of lake water into the ground water system (Figure 3). Infiltration of stream water is relatively minor and occurs only along short, local reaches of streams.

The ground water system discharges to streams and to Mirror Lake. Using baseflow separation methods developed by the Institute of Hydrology (1980a, 1980b) and by Nathan and McMahon (1990), Mau and Winter (1997) estimated that baseflow accounts for about 20 to 55% of streamflow. In the present study, baseflow is assumed to be 40% of streamflow over the long term. Table 1 gives long-term average streamflow (Rosenberry and Winter 1993) and estimated long-term average baseflow in Streams W, NW, and E at their inlets to Mirror Lake. Ground water enters Mirror Lake along the mud-free, near-shore lake bottom on the west, north, east, and a portion of the south sides of the lake (Asbury 1990). Minute amounts of ground water also enter Mirror Lake by upward seepage through the lacustrine mud, or directly from bedrock where it crops out on the east side of the lake. Rosenberry and Winter (1993) calculated that ground water discharge to Mirror Lake is approximately 80,000 m^3 year^{-1}. They indicated, however, that this value is an order-of-magnitude estimate.

Hydraulic heads in the Mirror Lake area are measured in 70 piezometers constructed in glacial deposits and in 31 wells drilled into bedrock (Figure 3). Depths of piezometer screen midpoints range from 1 to 40 m below land surface, and depths of bedrock wells range from 46 to 305 m below land surface. Due to limited drilling access at upper altitudes, all piezometers and wells except one have been constructed on lower hillsides and valleys, below an altitude of 360 masl. The exception is well FS7, drilled 60 m into the bedrock on the ridge near the highest point on the northwestern corner of the study area (Figure 1). Packers have been installed in many of the bedrock wells to allow multilevel monitoring (Hsieh et al. 1993). Long-term average hydraulic heads measured in piezometers and wells are summarized by Tiedeman et al. (1997). Head measurements made over several years in the 15 piezometer nests (Figure 3) show no evidence of perched zones of water in the glacial deposits. On lower hillsides and valleys, the water table lies within 20 m of land surface, and in many locations, less than 10 m below land surface. At the upper-altitude well FS7, the water table is about 15 m below land surface. This observation suggests that the water table remains shallow at all altitudes in the Mirror Lake area, from the low point at the Pemigewasset River to the ridge top near the northwestern corner of the study area.

Ground Water Flow Model

Ground water flow through glacial deposits and crystalline bedrock in the Mirror Lake area is simulated by the U. S. Geological Survey Modular Three-Dimensional Finite-Difference Ground Water Flow Model (McDonald and Harbaugh 1988), commonly known as MODFLOW. By using MODFLOW, we assume that a continuum approach is suitable at the scale of our study. The MODFLOW computer code is modified slightly so that drying of finite-difference cells is delayed until the end of the iterative solution loop. The Streamflow-Routing Package developed by Prudic (1989) is included in the model to simulate baseflow in streams and interaction between stream water and ground water. The steady-state model in this study simulates long-term average flow conditions. Annual, seasonal, and shorter-period variations are not considered.

Four simplifying assumptions are used in developing the ground water flow model:

1. Recharge from precipitation is assumed areally uniform. Although variations in precipitation, slope, vegetation, and soil could lead to variations in infiltration, studies of such spatial variability have not been carried out in the Mirror Lake area.
2. Riparian evapotranspiration along stream banks is neglected. This assumption is based on the work of Rutledge (1993), who analyzed streamflow in 166 surface water basins in the Appalachians and Piedmont area, and found that for each basin, approximately 10% of ground water discharge is consumed by riparian evapotranspiration. When riparian evapotranspiration is neglected, baseflow equals ground water discharge to streams less infiltration of stream water along losing stream reaches.
3. Flow in the bedrock is assumed to occur in the uppermost 150 m. Although permeable fractures are encountered at greater depths, results of packer tests suggest that most of the permeable fractures occur within 150 m below bedrock surface. By limiting ground water flow in the bedrock to the uppermost 150 m, the model ignores the possibility of deeper ground water flow at larger regional scales that extend beyond the Mirror Lake area.
4. The distribution of hydraulic conductivity is represented by dividing the model domain into large homogeneous, isotropic zones. In the bedrock, geologic complexity precludes delineating hydraulic conductivity zones according to rock type distribution. Highly conductive fracture clusters are not explicitly represented in the model, because hydraulic test results at the 100 m scale suggest that ground water flow is likely controlled by the less conductive fracture network, and because highly conductive fracture clusters have been characterized only in a small part of the study area. Because small-scale heterogeneities are neglected, the simulation characterizes the overall ground water flow system, but not local details.

The three-dimensional model domain extends horizontally over the Mirror Lake area (Figure 1) and vertically from the water table to 150 m below the bedrock surface. The horizontal dimensions are discretized into a grid of 89 rows by 85 columns of rectangular cells. The central part of the grid has square cells 17 m on each side. Outside the central part, the cell dimension increases towards the perimeter in two stages, from 17 m to 50 m, and then from 50 m to 150 m.

The vertical dimension is discretized into five model layers. The upper two model layers (layers 1 and 2) represent glacial deposits, Mirror Lake, lacustrine mud, and the Pemigewasset River (Figure 4). In the central part of the grid where the thickness of glacial deposits is known from drilling and geophysical surveys, layers 1 and 2 are each assigned half the thickness of the glacial deposits. Where the thickness of glacial deposits is unknown, layers 1 and 2 are each assigned a uniform thickness of 4.5 m. At bedrock outcrops, layers 1 and 2 are absent. The lower three model layers represent the bedrock. Each of layers 3, 4, and 5 has a uniform thickness that is 30 m, 60 m, and 60 m, respectively. The vertical discretization in the bedrock is admittedly coarse. However, because the model is intended for simulating the large-scale flow system, the relatively coarse vertical discretization is considered acceptable.

The vertical surfaces that bound the sides of the model domain are specified as no-flow boundaries. On the east, north, and west

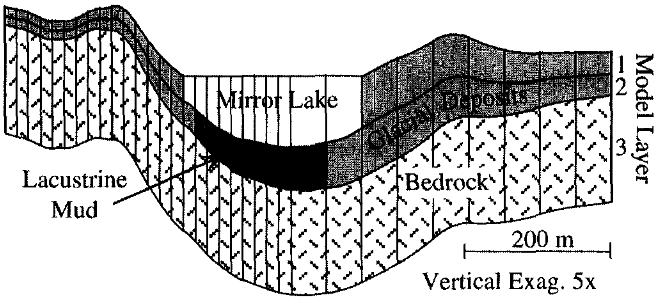

Figure 4. Vertical section along columns 40 to 72 of row 50 of the model grid showing glacial deposits, Mirror Lake, lacustrine mud, and bedrock in model layers 1, 2, and 3. Layer boundaries are shown schematically; true cell shapes are rectangular on a side.

sides, these vertical boundary surfaces approximately lie beneath the Pemigewasset River, Leeman's Brook, Paradise Brook, and a reach of Hubbard Brook (Figure 1), and the no-flow boundary condition assumes that ground water does not flow across the river and streams, but discharges into them. On the south side and near the northwest corner of the model domain, the vertical boundary surfaces lie beneath major topographic divides, and the no-flow boundary condition assumes that these divides are also ground water divides. Because these topographic divides lie along major ridges, they are thought to be likely locations where ground water and surface water basin boundaries coincide.

The top surface of the model domain is the water table, the position of which is computed during simulation. Model cells that lie above the water table become "inactive" and are eliminated from the simulation. Model cells that contain or lie below the water table remain "active" in the simulation. Recharge from precipitation is represented as an areally uniform flux applied to the uppermost active cell throughout the model domain, and is estimated by model calibration. The bottom surface of the model domain (150 m below bedrock surface) is specified as a no-flow boundary.

Streams in the model domain include Hubbard Brook, Norris Brook, Paradise Brook, Leeman's Brook, Streams W, NW, and E, and their tributaries. Streams are simulated by the Streamflow-Routing Package of MODFLOW (Prudic 1989), which considers streams to lie over model cells in layer 1. Flow from an underlying model cell to a stream reach is computed by (Prudic 1989, page 7):

$$Q_s = C_s(H_s - h) \qquad (1)$$

where C_s is the streambed conductance ($L^2\,T^{-1}$), H_s is the average stream stage in the reach (L), and h is the hydraulic head in the cell (L). If H_s is lower than h, ground water discharges into the stream (Q_s is negative). If H_s is higher than h, stream water infiltrates into the ground water system (Q_s is positive), providing there is streamflow in the reach. After Q_s is computed for a reach, it is added to the streamflow from the adjacent upstream reach, and the sum is routed to the adjacent downstream reach. In the present model, streambed conductance, C_s , is set to a large value for all reaches. It is likely that on the high-gradient streams in the Mirror Lake area, there is little deposition of fine sediments that would create resistance to ground water-stream interaction. This approach causes hydraulic head in the underlying cell to be close to the stream stage in areas of the model where the computed Q_s is nonzero.

Distribution of hydraulic conductivity in the model is represented by dividing the model domain into five homogeneous and

Table 1
Long-Term Average Streamflow, Long-Term Average Baseflow, and Simulated Baseflow in the W, NW, and E Streams at Their Inlets to Mirror Lake

Stream	Long-Term Average Streamflow ($m^3\,year^{-1}$)	Long-Term Average Baseflow[1] ($m^3\,year^{-1}$)	Simulated Baseflow ($m^3\,year^{-1}$)
W	157,000	63,000	62,800
NW	249,000	100,000	92,300
E	12,000	5000	9200
Total	418,000	168,000	164,300

[1]Estimated as 40% of streamflow.

isotropic zones: (1) glacial deposits, excluding sand and gravel south of Mirror Lake; (2) sand and gravel south of Mirror Lake; (3) bedrock beneath lower hillsides and valleys; (4) bedrock beneath upper hillsides and hilltops; and (5) lacustrine mud. Because the sand and gravel south of Mirror Lake exert an important control on the subsurface outseepage from the lake, this deposit is considered a separate zone from the rest of the glacial deposits. The division of the bedrock into two large hydraulic conductivity zones is necessary to simulate a water table that lies close to land surface over the entire Mirror Lake area. As shown by the land surface topography in Figure 1, upper hillsides (at altitudes greater than 360 masl) are significantly steeper than lower hillsides (at altitudes less than 360 masl). With uniform recharge from precipitation and homogeneous hydraulic conductivity in the bedrock, it is not possible to simulate a water table that exhibits an abrupt change in slope. Therefore, the bedrock is divided along the 360 masl topographic contour into two separate hydraulic conductivity zones. This division extends vertically through model layers 3, 4, and 5. The bedrock hydraulic conductivity under the upper hillsides is expected to be less than the bedrock hydraulic conductivity under the lower hillsides. A possible geologic explanation for the smaller hydraulic conductivity is that the rock beneath upper hillsides might be less fractured.

Model Calibration

Model calibration consists of estimating (1) recharge from precipitation, (2) hydraulic conductivity of the glacial deposits, and (3) hydraulic conductivity of the two bedrock zones. Hydraulic conductivities of the sand and gravel (south of Mirror Lake) and the lacustrine mud are assumed to be known, because they occupy relatively small parts of the model domain. Based on the values used by Rosenberry and Winter (1993), the hydraulic conductivity of the sand and gravel is fixed at 5×10^{-5} m s^{-1}, and the hydraulic conductivity of the lacustrine mud is fixed at 1×10^{-9} m s^{-1}.

Calibration data consist of (1) long-term average baseflow in Streams W, NW, and E given in Table 1, and (2) long-term average hydraulic heads in piezometers and bedrock wells given by Tiedeman et al. (1997). These two data sets are referred to as (1) observed baseflow and (2) observed hydraulic heads. The computer program MODFLOWP (Hill 1992) is used to determine optimal parameters that minimize the sum of squared weighted differences between observed and simulated quantities:

$$S(b) = \sum_{i=1}^{n_h} \{w_i^{1/2}[H_i - h_i(b)]\}^2 + \sum_{j=1}^{n_q} \{w_j^{1/2}[Q_j - q_j(b)]\}^2 \qquad (2)$$

where

b is the vector of parameters to be estimated

H_i is the observed hydraulic head at location i (L)

$h_i(b)$ is the simulated hydraulic head at location i computed with parameter vector b (L)

n_h is the number of hydraulic head observations (91 in this study)

$w_i^{1/2}$ is the weight on H_i (L^{-1})

Q_j is the observed baseflow in stream j ($L^3 T^{-1}$)

$q_j(b)$ is the simulated baseflow in stream j computed using parameter vector b ($L^3 T^{-1}$)

n_q is the number of streams (three in this study)

$w_j^{1/2}$ is the weight on Q_j ($T L^{-3}$).

In the actual regression, the common log of hydraulic conductivity is estimated; the result is then exponentiated to yield the estimate for hydraulic conductivity.

Because hydraulic head and baseflow are different quantities with different units, weighting is necessary to combine both into a single optimization criterion. In principle, the weights can be derived from the statistical structure of errors in hydraulic head and baseflow. In the present study, this statistical structure is unknown, and the weights $w^{1/2}$ are chosen to express the importance of the baseflow observations relative to the hydraulic head observations. The weight on each hydraulic head observation is set to 1 m^{-1}. For Streams W and NW, the weights are chosen such that a 1% difference between observed and simulated baseflow is equivalent to a 1 m difference between observed and simulated hydraulic head. For Stream E, the weight is such that a 5% difference between observed and simulated baseflow is equivalent to a 1 m difference between observed and simulated hydraulic head. The unequal baseflow weighting is appropriate because Stream E drains a much smaller subbasin than Streams W or NW, and therefore exerts a much smaller influence on the overall ground water flow system. The unequal weighting allows a larger tolerance when matching observed and simulated baseflow in Stream E than in Streams W and NW.

Results of model calibration are given in Table 2. The estimated 28 cm $year^{-1}$ of recharge from precipitation represents approximately 23% of the long-term average precipitation measured at the Forest Service station. This value is near the lower limit of the range (19 to 53 cm $year^{-1}$) estimated by Mau and Winter (1997). For glacial deposits excluding sand and gravel south of Mirror Lake, the estimated hydraulic conductivity of 1.7×10^{-6} m s^{-1} lies between the slug-test estimates for till and for sand and gravel as determined by Winter and Rosenberry (1995). For bedrock beneath lower hillsides and valleys, the estimated hydraulic conductivity of 3.2×10^{-7} m s^{-1} is consistent with the effective hydraulic conductivity of 2×10^{-7} m s^{-1} determined by Hsieh and Shapiro (1994) for a 100 m size block of bedrock on a lower hillside. For bedrock beneath upper hillsides and hilltops, the hydraulic conductivity estimate is 6.3×10^{-8} m s^{-1}. Because this value is based largely on the single hydraulic head measured at the upper-altitude well FS7, it should be considered preliminary and requires further confirmation with additional data.

Uncertainty in the estimated parameters is characterized by linear, individual, 95% confidence intervals (Table 2) calculated by the procedure described by Hill (1992). The size of each confidence interval in Table 2 is small relative to the corresponding optimal parameter estimate, suggesting that the optimal estimates are well constrained. However, these confidence intervals are computed under the assumption that model attributes such as geometry, zonation, and boundary conditions are perfectly known. Errors or uncertainties in these model attributes could increase the uncertainty of the estimated parameters.

On an overall level, there is relatively good match between simulated and observed quantities. Table 1 shows that simulated and observed baseflows differ by less than 10% in Streams W and NW, which together account for 90% of total baseflow. A larger percentage discrepancy occurs in Stream E due to its smaller baseflow and lower weight. The standard error of hydraulic head, which is the average (root-mean-square) weighted discrepancy between the simulated and observed hydraulic heads, is 2.7. This standard error is small compared to the 445 m of elevation difference between the lowest and highest observed hydraulic head. At FS7 (Figure 1), the upper-altitude well, the difference between simulated and observed heads is less than 1 m.

On a detailed level, however, there is distinct lack of fit between observed and simulated hydraulic heads. Figure 5 shows hydraulic-head contours in the vertical section AB (Figure 3) on a hillside west of Mirror Lake. Over the entire section, simulated contours match observed contours in conveying the sense of lateral ground water flow in the downslope direction. However, on a detailed level, the two sets of contours show distinct differences. The observed contours indicate that the vertical hydraulic gradient varies significantly along the section, suggesting that ground water flow could be affected by local heterogeneities in areal recharge

Table 2
Optimal Value and Linear, Individual, 95% Confidence Interval for Estimated Parameters

Estimated Parameter	Optimal Value	95% Confidence Interval
Recharge from precipitation (cm $year^{-1}$)	28	26 to 30
Hydraulic conductivity of glacial deposits excluding sand and gravel south of Mirror Lake (m s^{-1})	1.7×10^{-6}	1.6×10^{-6} to 1.9×10^{-6}
Hydraulic conductivity of bedrock beneath lower hillsides and valleys (m s^{-1})	3.2×10^{-7}	2.8×10^{-7} to 3.7×10^{-7}
Hydraulic conductivity of bedrock beneath upper hillsides and hilltops (m s^{-1})	6.3×10^{-8}	5.7×10^{-8} to 6.9×10^{-8}

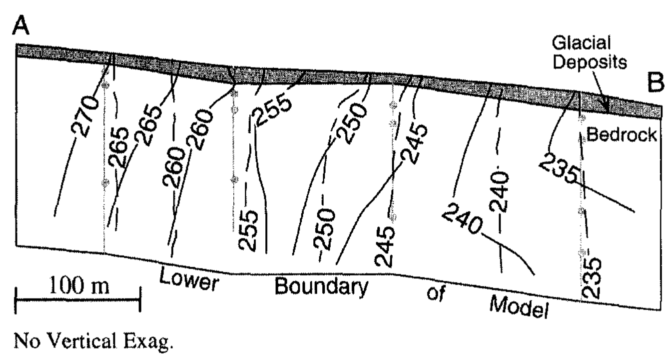

Figure 5. Long-term average observed (solid contours) and simulated (dashed contours) hydraulic head (in masl) along vertical section A-B (see Figure 3 for location of section line). Gray lines are bedrock wells; gray circles show hydraulic-head measurement locations.

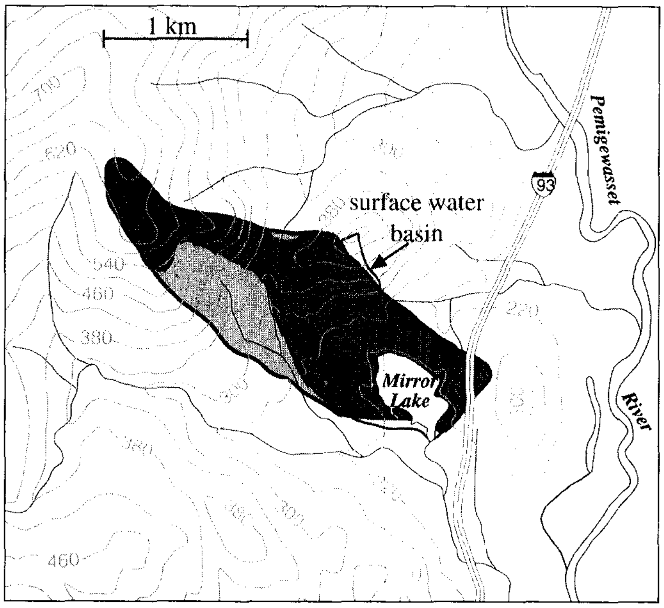

Figure 6. Horizontal extent of the Mirror Lake ground water basin. Thick solid line indicates the lateral boundary of the basin, heavily shaded area is the recharge zone, lightly shaded area is the underflow zone, thick dashed line is the boundary between the two zones.

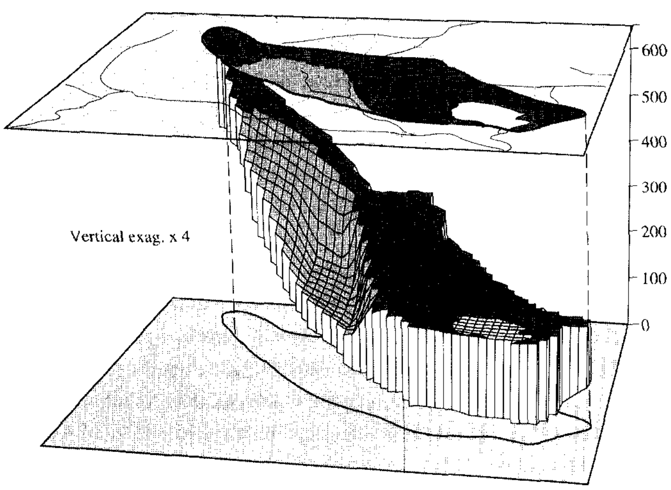

Figure 7. Three-dimensional rendering of the Mirror Lake ground water basin.

and/or hydraulic conductivity. For example, Harte (1992) simulated ground water flow along a cross section from the northwest corner of the study area to Mirror Lake, and showed that including highly permeable kame terraces on the hillsides resulted in variations in the simulated vertical hydraulic gradient in both the glacial deposits and the bedrock. In contrast to the observed contours in Figure 5, the simulated contours in our model are close to a regular pattern, because the model does not include local heterogeneities in recharge and hydraulic conductivity. This comparison shows that the model results in this study should be viewed as a depiction of the overall ground water flow system, not of local details.

Delineation of Ground Water Basin

With a calibrated ground water flow model, the three-dimensional extent of the Mirror Lake ground water basin can be determined by tracking ground water flowpaths. The computer program MODPATH (Pollock 1994), a particle-tracking post-processor package for MODFLOW, is used to track flowpaths in a reverse direction from Mirror Lake and its inlet streams back to the water table. The volume of subsurface spanned by the flowpaths is taken to be the ground water basin. Figures 6 and 7, respectively, show the horizontal and three-dimensional extent of the basin. Note that the region of subsurface outseepage from the south side of Mirror Lake is not included in the Mirror Lake ground water basin, because ground water in this region flows towards Hubbard Brook.

Figures 6 and 7 suggest that a ground water basin can take on a complex shape. In Figure 6, the horizontal extent of the Mirror Lake ground water basin is divided into a "recharge zone" (heavily shaded area) and two "underflow zones" (lightly shaded areas). The recharge zone illustrates the part of the ground water basin that extends vertically from the basin bottom up to the water table. Recharge entering this zone eventually drains to Mirror Lake or its inlet streams. The two underflow zones illustrate parts of the ground water basin that do not extend vertically up to the water table. These are areas where the Mirror Lake ground water basin lies beneath adjacent ground water basins. In Figure 7, the recharge zone corresponds to the heavily shaded surface, which is the water table, and the larger underflow zone corresponds to the lightly shaded surface. The smaller underflow zone is hidden from view.

A comparison between the recharge zone and the surface water basin suggests that Mirror Lake and its inlet streams drain a ground water recharge area that is about 1.5 times the area of the surface water basin. The ground water basin extends far up the hillside on the northwestern part of the study area. Therefore, ground water flows across the topographic divide that forms the northwestern boundary of the surface water basin. Simulated flowpaths suggest that a portion of recharge to the northwestern part of the study area flows at depth beneath Norris Brook to discharge into Mirror Lake or its inlet streams. In the shallow saturated zone around Norris Brook, ground water flows to Norris Brook to form a separate ground water basin. Therefore, a part of the Mirror Lake ground water basin lies beneath the adjacent ground water basin that drains into Norris Brook.

Ground Water Flow in Glacial Deposits and Bedrock

The partitioning of ground water flow through glacial deposits and bedrock can be accomplished by calculating a ground water budget. In this study, ground water budget components are calculated by the computer program ZONEBUDGET (Harbaugh 1990) using simulation results from MODFLOW. Uncertainty in the budget components is assessed by calculating approximate, individual, 95% confidence intervals following the procedure described by Hill (1994). The present budget does not include evapotranspiration, surface runoff that moves through the shallow subsurface to streams, or lake evaporation, because these processes are not simulated by the ground water flow model.

Depending on the position of the water table, recharge from precipitation can enter the ground water system in the glacial deposits or in the bedrock. Figure 8 shows the simulated water table in the Mirror Lake area. The shaded areas indicate where the simulated water table lies in glacial deposits, and therefore, where recharge enters the ground water system from the glacial deposits. The unshaded areas indicate where the simulated water table lies in bedrock. In these areas, recharge reaches the water table either by infiltration through the overlying unsaturated glacial deposits, or directly from land surface where bedrock crops out. For budget calculations, recharge to the ground water basin includes:

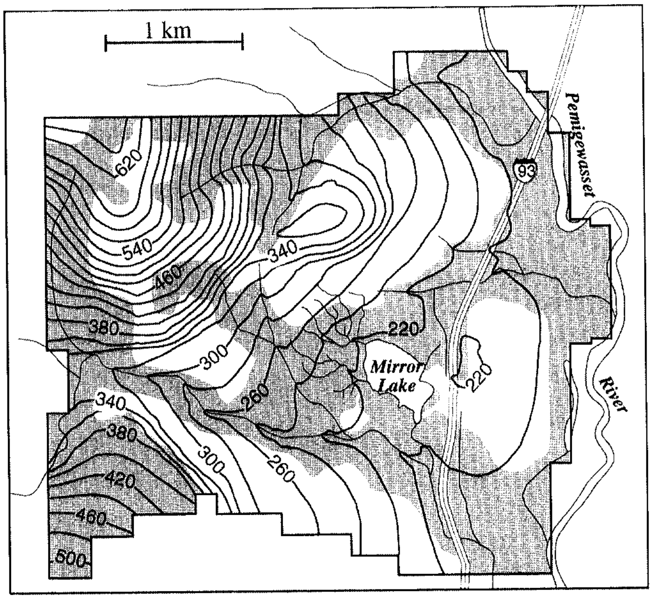

Figure 8. Simulated water table (in masl). Shaded regions indicate areas where the water table lies in glacial deposits. Unshaded regions indicate areas where the water table lies in bedrock.

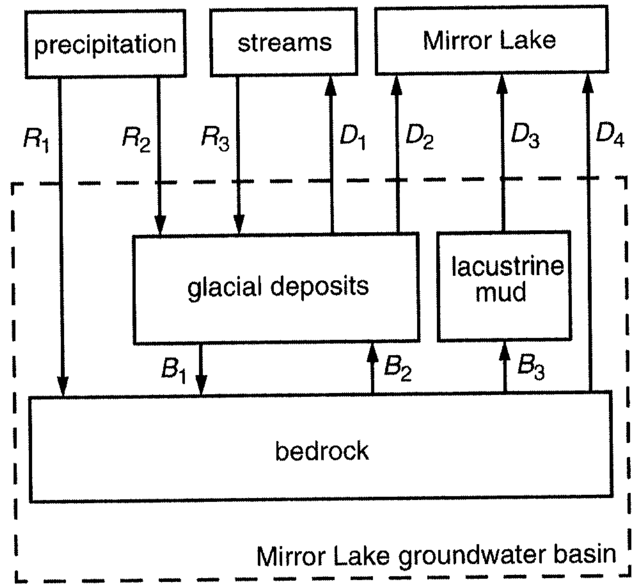

Figure 9. Components of the ground water budget for the Mirror Lake ground water basin. The boxes labeled "precipitation," "streams," and "Mirror Lake" represent sources of ground water recharge to the basin and sinks for ground water discharge from the basin.

R_1 = recharge from precipitation to bedrock ($L^3\ T^{-1}$),
R_2 = recharge from precipitation to glacial deposits ($L^3\ T^{-1}$)
R_3 = recharge from streams to glacial deposits ($L^3\ T^{-1}$), which occurs along losing reaches of W and NW streams.

Flow between hydrogeologic units in the ground water basin includes:
B_1 = flow from glacial deposits to bedrock ($L^3\ T^{-1}$)
B_2 = flow from bedrock to glacial deposits ($L^3\ T^{-1}$)
B_3 = flow from bedrock to lacustrine mud ($L^3\ T^{-1}$).

Discharge from the ground water basin includes:
D_1 = discharge from glacial deposits to streams ($L^3\ T^{-1}$),
D_2 = discharge from glacial deposits to Mirror Lake ($L^3\ T^{-1}$),
D_3 = discharge from lacustrine mud to Mirror Lake ($L^3\ T^{-1}$), and
D_4 = discharge from bedrock to Mirror Lake ($L^3\ T^{-1}$), which occurs where bedrock crops out on the east side of the lake.

The budget components are illustrated in Figure 9. Under steady state, total recharge to the ground water basin, defined as

$$R_T = R_1 + R_2 + R_3 \tag{3}$$

equals total discharge from the basin, defined as

$$D_T = D_1 + D_2 + D_3 + D_4 \tag{4}$$

Results of budget calculations (Table 3) suggest that a total recharge (R_T) of approximately 300,000 m^3 $year^{-1}$ enters the Mirror Lake ground water basin. Nearly all the recharge is derived from precipitation ($R_1 + R_2$); only a small portion is derived from infiltration of stream water along losing reaches of streams (R_3). Of the recharge from precipitation, about half enters the basin in areas where the simulated water table lies in bedrock; the other half enters the basin in areas where the simulated water table lies in glacial deposits.

Within the Mirror Lake ground water basin, flows between hydrogeologic units are strongly controlled by the direction of the vertical flow component. Flow from glacial deposits to bedrock (B_1) occurs primarily under upper hillsides, where the vertical flow component is predominantly downward. Because significant parts of the glacial deposits on upper hillsides are unsaturated, flow from glacial drift to bedrock is relatively small. By contrast, flow from bedrock to glacial deposits (B_2) occurs primarily under lower hillsides and valleys, where the vertical flow component is predominantly upward. Because little outflow from the bedrock can exit through the lacustrine mud or the outcrop on the east side of Mirror Lake, nearly all outflow from the bedrock exits into the overlying glacial drift.

Of the total discharge from the Mirror Lake ground water basin (D_T), slightly more than half flows into Streams W, NW, and E (D_1), slightly less than half flows into Mirror Lake from the glacial deposits (D_2), and minute amounts enter Mirror Lake from the lacustrine mud (D_3) and from bedrock outcrops on the east side of the lake (D_4). Total seepage of ground water to Mirror Lake ($D_2 + D_3 + D_4$) is approximately 130,000 m^3 $year^{-1}$. Although this value is higher than Rosenberry and Winter's (1993) estimate of 80,000 m^3 $year^{-1}$, it is within the range of uncertainty for their estimate. The discrepancy between the seepage estimates is due to differences in the hydraulic conductivities and hydraulic gradients used in each calculation.

The size of each confidence interval (Table 3) is small in comparison to the corresponding budget component, suggesting that the computed budget components are well constrained. However, the caveats on confidence intervals for model parameters equally apply to confidence intervals on simulated quantities. Because flow from the lacustrine mud to Mirror Lake (D_3) and flow from bedrock to Mirror Lake (D_4) are both minute, their confidence intervals are not calculated because they are probably not meaningful.

The ground water budget in Table 3 provides insight into flow through a ground water basin in glacial deposits and bedrock. Of the total recharge to the basin, a portion travels along flowpaths that stay exclusively in glacial deposits, while the remainder travels along flowpaths that involve movement in bedrock. These two types of

Table 3
Ground Water Budget of Mirror Lake Ground Water Basin, and Linear, Individual, 95% Confidence Intervals for Budget Components

Budget Component	Computed Value (m^3 $year^{-1}$)	95% Confidence Interval (m^3 $year^{-1}$)
Recharge to ground water basin		
Precipitation to bedrock[1] (R_1)	147,000	136,000 to 158,000
Precipitation to glacial deposits[2] (R_2)	145,000	136,000 to 154,000
Streams to glacial deposits[3] (R_3)	5000	4000 to 6000
Total Recharge ($R_T = R_1 + R_2 + R_3$)	297,000	Not computed
Flow between hydrogeologic units in ground water basin		
Glacial deposits to bedrock (B_1)	34,000	29,000 to 39,000
Bedrock to glacial deposits (B_2)	178,000	165,000 to 191,000
Bedrock to lacustrine mud (B_3)	1000	Not computed
Discharge from ground water basin		
Glacial deposits to streams (D_1)	169,000	159,000 to 179,000
Glacial deposits to Mirror Lake (D_2)	125,000	118,000 to 132,000
Lacustrine mud to Mirror Lake (D_3)	1000	Not computed
Bedrock to Mirror Lake[4] (D_4)	2000	Not computed
Total Discharge ($D_T = D_1 + D_2 + D_3 + D_4$)	297,000	Not computed

[1]In areas where simulated water table lies in bedrock.
[2]In areas where simulated water tables lies in glacial deposits.
[3]Along losing reaches of Streams W and NW.
[4]Where bedrock crops out at east side of Mirror Lake.

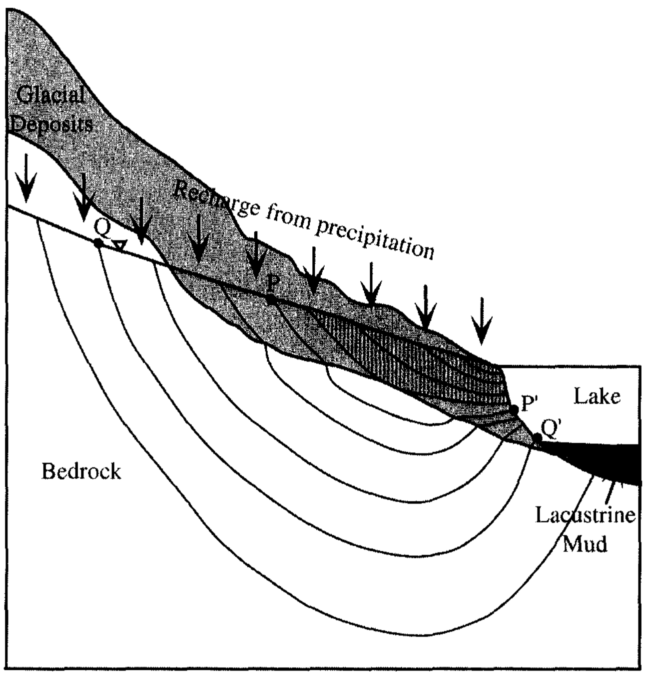

Figure 10. Schematic vertical section illustrating flowpaths through glacial deposits and bedrock.

flowpaths are conceptually illustrated in Figure 10. The stippled region shows flowpaths that stay exclusively in glacial deposits. Outside the stippled region, flowpaths involve movement in bedrock. Note, however, that the travel distance in bedrock varies from relatively short (for example, flowpath P-P′) to relatively long (for example, flowpath Q-Q′). As shown in Figure 10, the flow that involves movement in bedrock is simply the total inflow to bedrock, or ($R_1 + B_1$). Conversely, the flow that stays exclusively in glacial deposits is the total recharge minus flow that involves movement in bedrock, or ($R_T - R_1 - B_1$). Applying this concept to the Mirror Lake area suggests that a flow of approximately 181,000 m^3 $year^{-1}$ moves through the bedrock. This value is approximately 60% of total recharge to the ground water basin. The flow that stays exclusively in glacial deposits is about 116,000 m^3 $year^{-1}$. This value is approximately 40% of total recharge to the ground water basin.

Discussion

The results of this study support the premise that a ground water basin does not always coincide with the volume obtained by extending a surface water basin vertically downward to the bedrock surface. Instead, a ground water basin is more appropriately delineated by tracking ground water flowpaths with a calibrated, three-dimensional ground water flow model. In general, the model must cover an area that is substantially larger than that of the surface water basin. In this study, the modeled area is about 10 times the area of the surface water basin. The ground water basin determined by flowpath tracking could take on a complicated, three-dimensional shape, possibly with portions extending beneath adjacent ground water basins. The area of recharge to the ground water basin could differ substantially from the area of the surface water basin. In this study, the simulated ground water recharge area for Mirror Lake and its inlet streams is about 1.5 times larger than the area of the surface water basin. Therefore, if the extent of the ground water flow model were limited to the surface water basin, the calibrated recharge rate would be about 1.5 times larger, because a higher recharge would be needed over a smaller area to match the stream baseflows.

In analyzing ground water flow through glacial deposits and bedrock, it is useful to divide flowpaths into two categories: (1) flowpaths that remain exclusively in glacial deposits; and (2) flowpaths that involve movement through bedrock. This separation is important for calculations of the chemical budget, because ground water that contacts the bedrock may contain different amounts of dissolved chemicals than ground water that flows only through the glacial deposits. The present study suggests that the partitioning of flow between glacial deposits and bedrock is strongly controlled by (1) the transmissivity (product of hydraulic conductivity and saturated thickness) of glacial deposits versus bedrock, and (2) the amount of recharge to glacial deposits versus bedrock. In the Mirror Lake area, the hydraulic conductivity of the glacial deposits is about an order of magnitude greater than that of the bedrock, but the saturated thickness of the bedrock is about an order of magnitude greater than that of the glacial deposits. Therefore, the transmissivity of the two hydrogeologic units is about the same, meaning that both units can transmit similar amounts of lateral flow under similar hydraulic gradients. In addition, simulated recharge from precipitation to glacial deposits is about the same as simulated recharge from precipitation to bedrock. The consequence of these two factors is that about 60% of flow in the Mirror Lake ground water basin travels along flowpaths that involve movement in bedrock, while the remaining 40% travels solely in glacial deposits.

Although the ground water flow model developed for the Mirror Lake area is admittedly simple, model calibration yields parameter estimates that are consistent with findings from field tests and other analyses, and there is a good overall fit between simulated and observed quantities. In addition, Tiedeman et al. (1997) investigated alternative parameterization schemes and found that neither the overall model fit nor the local details were improved by introducing a 10-to-1 ratio of horizontal-to-vertical anisotropy in the hydraulic conductivity of the glacial deposits, or by allowing hydraulic conductivity to vary with depth in the modeled portion (uppermost 150 m) of the bedrock. Taken together, these findings suggest that, for the purpose of this study, the glacial deposit and fractured bedrock system can be represented in a standard ground water flow model by dividing the model domain into several large, homogeneous zones. Although the model might fail to accurately simulate the local details of ground water movement, it does give a reasonable depiction of the overall ground water flow system and budget.

Because only one measurement of hydraulic head is available for upper altitude regions (higher than 360 masl) in the Mirror Lake area, simulation results for these regions should be considered preliminary. The steepening of the simulated water table at upper altitudes is accomplished in the present study by using a lower hydraulic conductivity for bedrock beneath upper hillsides and hilltops, while maintaining a uniform recharge from precipitation over the entire modeled area. Additional measurements of hydraulic head at upper altitude regions of the study area may serve to refine the model in these regions. However, the shape of the Mirror Lake ground water basin is not highly sensitive to the hydraulic conductivity of the bedrock beneath the upper hillsides and hillslopes. Tiedeman et al. (1997) found that when the upper altitude hydraulic head measurement is not used in the regression, and a uniform hydraulic conductivity is estimated for the entire volume of bedrock in the model, the horizontal extent of the ground water basin is similar to that shown in Figure 6.

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